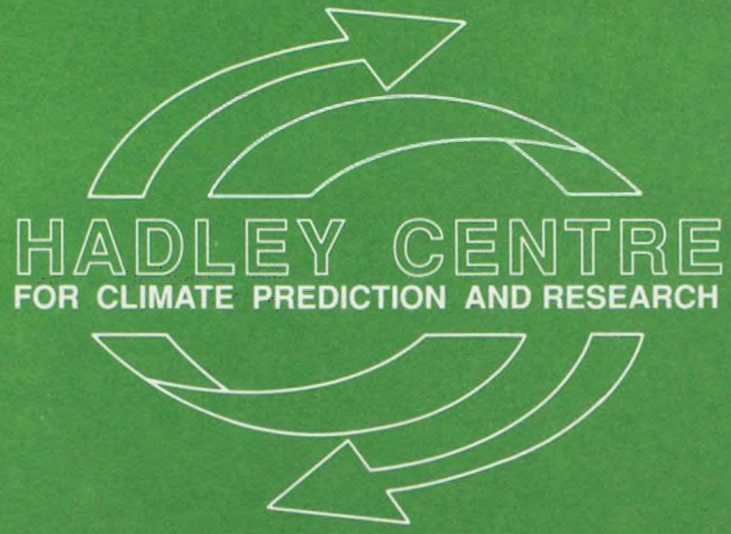

The oceanic carbon cycle

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CRTN 19

September 1991

CLIMATE RESEARCH TECHNICAL NOTE

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N K TAYLOR

Hadley Centre for Climate Prediction and Research
Meteorological Office
London Road
Bracknell
Berkshire RG12 2SY
U.K.

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1. Introduction

In recent years there has been increasing interest in the response of climate to increasing emissions of greenhouse gases, most notably carbon dioxide, caused by man's activities. Combustion of fossil fuels and deforestation has led to a 26% increase in atmospheric CO_2 concentration since the industrial revolution. The rate and nature of climate change will depend upon the rate of increase of atmospheric CO_2, which is dependent upon both the rate of future emissions and the fraction of these emissions that remains airborne.

The oceans play a key role in the earth's climate system. Because of their great heat capacity, they even out the daily, seasonal and spatial differences in the thermal energy received from the sun. Just 2.5m depth of water has the same heat capacity per unit area as the whole depth of the atmosphere. The ocean also has a more subtle, yet potentially important, role in climate through its effect on the global carbon cycle. It is by far the largest reservoir of the worlds' mobile carbon (Figure 1) and it is widely accepted that it controls the atmospheric CO_2 concentration and its variations with time. It is also an important reservoir for taking up anthropogenic CO_2. At present only about half of the emitted CO_2 stays in the atmosphere. The remainder is absorbed in the oceans and by the terrestrial biota, although details of the partitioning are not yet well known. Recent data suggest that the ocean is not such an important sink for anthropogenic CO_2 as was previously supposed - the current estimate of its' sink strength is between 0.6 and 1.4 GtC (gigatonnes of carbon) per year (Tans *et al*, 1990). This figure should be compared with the annual release of CO_2 from fossil fuel burning of 5.3 GtC. However, it is not just the uptake of anthropogenic CO_2 that should be considered, but also any potential changes to

the 'natural' carbon cycle which could be induced by climate change. Oceanic uptake of CO_2 is controlled by a complicated interaction between inorganic chemistry, ocean biology and by various patterns of mixing and circulation. It can thus be expected to react in a complex way to any changes in future climate.

Most of the oceans' carbon is in the form of dissolved inorganic carbon. At any one time the standing stock of living organic matter is relatively small, but it has a high rate of turnover - primary production in the surface layer of the sea is some 30-40% of the annual production of terrestrial vegetation. In temperate and higher latitudes marine primary production exhibits a strong seasonal cycle, with peaks in activity occuring during spring and summer in each hemisphere. The stirring up of nutrient-rich deep water in winter followed by the stabilization of the water column due to surface warming and reduced turbulence in spring, and increased sunlight, causes an explosive growth of phytoplankton populations. Satellite pictures of ocean colour have demonstrated the large geographical extent of this spring bloom, which occurs over wide swathes of the open ocean. Most marine primary production is recycled in the surface layers with only a relatively small portion (approx 10%) being exported out of the top few hundred metres or so as particulate organic carbon, which sinks and decomposes in deeper layers, or is incorporated into sediments.

The dynamics of the carbon cycle are linked to global cycles of certain other elements, particularly nitrogen and phosphorus. Rates of primary production over much of the ocean are limited by the supply of these nutrients.

2. Transfer Of CO_2 At The Air-Sea Interface And The Solubility Pump

For the purposes of examining the exchanges of CO_2 between the atmosphere and ocean we can divide the ocean into three functional layers. The surface layer reaches down to a depth of about 75m on average and is relatively well-mixed. Beneath this layer, extending to a depth of about 1km, is a relatively stagnant region (the thermocline) that is stabilized by decreasing temperature and increasing density. Below the thermocline, isolated from the surface waters, lies the deep ocean. However, under certain circumstances when the surface waters are sufficiently cold and/or saline, the thermocline weakens and the surface waters sink (or mix) to various depths and spread out horizontally (more accurately, along surfaces of equal density, called isopycnal surfaces). This produces a complex worldwide circulation pattern involving descending waters mostly in polar regions and upwelling of deep water from various depths elsewhere.

Surface waters exchange about 9% of their carbon with the atmosphere each year but the capacity of this relatively small, well-mixed, volume to take up additional CO_2 is quite limited. It is the transfer of excess CO_2 to the deep oceans by advection and mixing along isopycnal layers and by diffusive mixing across isopycnal surfaces that is the most important physical process determining the oceanic response to increasing CO_2 concentrations. These same processes will also be responsible for transferring heat away from the ocean surface during the course of a climatic warming, but they are not at present well understood.

The net flux of CO_2 into (or out of) the ocean at a particular location can be expressed as:

$$\text{Flux} = F_g = k\ S\ \Delta pCO_2$$

where S is the solubility, k is a gas transfer velocity, and ΔpCO_2 is the difference in CO_2 partial pressures betweeen sea water and air. Gas solubility varies with temperature so that more gas is dissolved at colder temperatures. This means that CO_2 leaves the ocean in the tropics and enters it at temperate and higher latitudes where the surface waters are cold (Figure 2). The regions where ΔpCO_2 is largest correspond to the regions where the mixed layer is deep in winter - a fact that is of some importance in the removal of CO_2 into the deep ocean.

Figure 3 is a schematic of the meridional circulation in the Atlantic Ocean. Deep water formation occurs episodically in localised regions in the North Atlantic and Antarctic oceans in winter. In the North Atlantic poleward-flowing warm surface water cools and becomes more saline (due to evaporation and brine rejection during ice formation), causing the water column to overturn convectively. A general flow of this bottom water towards the equator occurs at depth with the sinking at high latitudes being balanced by a slow upwelling in low latitudes. The Atlantic Deep Water formed in Northern high latitudes also eventually upwells to the surface in the region of the Antarctic continent where it helps to provide preconditioning for the formation of Antarctic Bottom Water, an essential factor in which is brine rejection during periods of ice formation in the Weddell and Ross Seas. When the surface distribution of pCO_2 is taken into consideration it is apparent that the deep, equatorward-flowing waters will be relatively carbon-rich. This deep circulation is extremely slow, returning subducted water back to the surface on a time scale of several hundred years. Thus CO_2 entering the deepest regions of the sea is removed from contact with the atmosphere for centuries. The sinking of water in polar regions, taking CO_2

with it, followed by flow towards the equator along the bottom with upwelling and degassing in equatorial regions is often called the "Solubility Pump".

The area of sea over which deep water formation occurs is limited both in space and time. Another process which takes place over a wider area is the ventilation of the permanent thermocline (Figure 4) . Ventilation is achieved by: a) convergence within the subtropical gyres and vertical diffusion between the thermocline and surface waters overlying it; and b) direct contact between the sea surface and isopycnal surfaces where they outcrop at the polar fronts. Since surface water can move downward along surfaces of constant density without working against gravity, the isopycnal outcrops are 'windows' by means of which water marked by atmospheric interaction can reach great depths in the oceans. CO_2 enters these layers, is mixed and advected along isopycnal surfaces, and is subducted below the surface into the thermocline. Process a) is relatively slow due to the inhibiting effects of stratification on vertical mixing. Carbon-14 measurements suggest that, on average, it takes hundreds to thousands of years for surface water to penetrate to well below the mixed layer of the major oceans. Process b) is much more rapid. Tritium observations in the North Atlantic, when interpreted in terms of time scales of exchange at the surface, give averages of 7-20 years for entire density surfaces in the upper and mid-depth subtropical thermocline. This is similar to the timescale over which climate change simulations are required, emphasising the need for numerical models capable of resolving thermocline ventilation mechanisms in some detail. Such a timescale also implies that CO_2 entering the thermocline will be returned back to the surface in a matter of decades, so it is the higher latitudes where surface water is exchanged with the cold water below the thermocline that is of significance to the long-term sequestration of CO_2.

3. The Biological Pump

Atmospheric CO_2 dissolves in seawater and is hydrated to form carbonic acid, H_2CO_3. This loses first one proton (H^+), then another, to form bicarbonate (HCO_3^-) and carbonate (CO_3^{--}) ions. The relative concentrations of these three inorganic species can be described by chemical equilibrium equations:

$$[H^+]\,[HCO_3^-] = K_1\,[H_2CO_3]$$

$$[H^+]\,[CO_3^{--}] = K_2\,[HCO_3^-]$$

where [ ] denotes concentrations and K_1, K_2 are the apparent equilibrium constants. The partial pressure exerted by the gaseous form is

$$pCO_2 = [H_2CO_3]/a$$

where a is solubility of gaseous carbon dioxide in seawater. The ability of carbon to exist as ions having different charges (as monovalent HCO_3^- and divalent CO_3^{--}) creates a chemical buffer system which regulates the pH and pCO_2 of the ocean.

To determine the carbon composition of seawater oceanographers need to measure just two quantities: a) the total concentration of inorganic carbon (ΣCO_2), defined as the sum of the concentrations of the three inorganic species; and b) the titration alkalinity (ALK), which is the amount of acid required to neutralise the weak bases present in seawater. Since by far the greatest contribution to alkalinity is provided by carbon species we can simplify the definition of alkalinity by considering carbonate alkalinity alone:

$$ALK = [HCO_3^-] + 2[CO_3^{--}]$$

The definition of alkalinity reflects conservation of charge in the ocean. When more negative charge is needed to balance the cations present in seawater, HCO_3^- is converted to $CO_3^{=}$; when less is needed, $CO_3^{=}$ is converted to HCO_3^-. Provided the pressure, temperature and salinity are known, the tendency for a volume of seawater to lose or take up CO_2 can be determined from measured values of ΣCO_2 and ALK, and expressed as the difference in CO_2 partial pressure between ocean and atmosphere, ΔpCO_2. pCO_2 increases approximately exponentially with ΣCO_2. It also decreases with increasing ALK. The sensitivity of pCO_2 to ΣCO_2 in seawater means that a given fractional change in pCO_2 produces a much smaller fractional change in ΣCO_2.

Concentrations of ΣCO_2 are found to be lower by some 10-15% in the surface layers than in the deep sea. The vertical distribution of alkalinity is similar, but with a rather smaller variation (Figure 5). It is photosynthetic activity by phytoplankton over the top 100m or so that maintains the surface depletion of ΣCO_2. CO_2 is taken up by phytoplankton during photosynthesis, and released during respiration by bacteria, phytoplankton, zooplankton and higher organisms. Most of the atmospheric carbon fixed by photosynthesis is respired in the surface ocean within a relatively short time (of the order of days to months). Approximately 10%, however, falls below the thermocline in the form of dead plant cells, zooplankton faecal pellets, carbonate shells and other biotic material, to be oxidised in the deep ocean. Production of carbonate shells by organisms such as Coccoliths uses up carbonate ions and so depletes alkalinity in the surface layers. Photosynthesis also causes a rise in alkalinity by the simultaneous consumption of H^+ ions but this is more than overcome by the loss of dissolved carbonate through shell formation. Thus ΣCO_2, ALK and nutrients are cast downwards by this continuous rain of particles of dead organic matter (organic detritus) and carbonate shells. A very small portion, representing about 5% of the downward flux, is incorporated into sediments on the ocean floor. The net

effect is that both ALK and ΣCO_2 are depleted in surface waters relative to deeper layers in most regions of the worlds oceans. If the surface water deficiency in ΣCO_2 were absent the pCO_2 in the atmosphere would be substantially increased

Organic carbon is recycled many times in the surface layer before it falls out as organic detritus. Most organic detritus is decomposed in the first 1000m, resulting in an O_2 minimum and a maximum in the ΣCO_2 profile at intermediate depths. Inorganic (carbonate) detritus is redissolved at greater depths, delivering carbon and alkalinity to the deep ocean. This cycling of carbon between the surface and the deep sea caused by the consumption of CO_2 by primary production at the surface and the enrichment of deep water in CO_2 from the decomposition and dissolution of detritus falling from the surface layers is termed the "Biological Pump".

The biological pump has an organic flux component ('soft tissue pump') and a carbonate flux component ('$CaCO_3$ counter pump'). Carbon exported below the thermocline cannot exchange with the atmosphere until those waters are returned back to the surface, a process that could take centuries. In this way carbon is effectively removed from the surface layers and surface pCO_2 is reduced. On the other hand, the downward flux of calcium carbonate in plankton shells (due to sinking), whilst also reducing ΣCO_2, leads to a decrease in surface alkalinity and thus an **increase** in surface pCO_2. This is because, as carbonate ions are removed from solution (by formation of solid $CaCO_3$), bicarbonate ions undergo a disproportionate reaction:

$$2HCO_3^- \quad ====> \quad H_2CO_3 + CO_3^{=}$$

to create carbonic acid. Thus the calcium carbonate component of the biological pump tends to act against the organic component.

Photosynthesis is dependent upon the availability of nutrients and light, and is often limited by lack of nitrogen and phosphorus. This is evidenced by the low concentrations of nitrate and phosphate found in temperate and tropical surface waters, except in regions of intense upwelling. The rate at which organic carbon and calcium carbonate are exported downwards from the surface waters is therefore probably determined in most locations by the rate at which nutrients are returned to the surface by upward circulation and mixing of water. However, at certain locations, where vertical mixing is rather fast, significant surface concentrations of nitrates and phosphates are found, especially in the Antarctic Ocean where the phosphate concentration is about 70% of its deep-water value; in warm surface water it is typically about 10%. Thus in the Antarctic Ocean productivity is not limited by nitrates or phosphates, but by some other factor, possibly light levels or by a shortage of iron (Martin *et al*, 1990). If the unused supplies of N and P could be taken up so that the biological pump ran at a higher rate of efficiency then pCO_2 in these regions could be reduced still further. Such a mechanism has been invoked to explain changes in pCO_2 that occurred during the last ice age (see e.g. Heinze *et al*, 1990), although there is as yet little evidence that this actually happened.

Measurements of gas bubbles trapped in Antarctic ice cores show a remarkable correlation between atmospheric CO_2 levels and temperature over the last 160,000 years (Barnola *et al*, 1987) (Figure 6). The CO_2 concentration during the last glacial period was about two-thirds of that in post-glacial times (Neftel *et al*, 1982). Furthermore, the changes in CO_2 associated with the close of glacial time occurred relatively rapidly over a period of less than 1,000 years. Analyses of Greenland ice cores has shown that during the last ice age the atmospheric CO_2

level varied between 200 and 260 ppm in parallel with climatic variations over timescales of the order of a few centuries (Stauffer *et al*, 1983). A number of models have been put forward which seek to explain the ice core data in terms of a more efficient utilization of nutrients at high latitudes during glacial times (Sarmiento and Toggweiler, 1984; Siegenthaler and Wenk, 1984; Knox and McElroy, 1984). In these 'Polar Nutrient' scenarios the ocean biota acts so as to amplify the effect of changes in physical characteristics, emphasising the important influence of both the polar seas and ocean biology on the CO_2 content of the atmosphere. Understanding past variations in climate may provide the key to predicting future climate changes.

4. Feedbacks

From the foregoing discussion it is clear that the physical, chemical and biological processes that occur within the ocean on the large space and time scales relevant to climate are highly interactive and cannot be considered in isolation when trying to understand the effects of increased atmospheric CO_2 concentrations on the ocean carbon cycle. Figure 7 illustrates the three carbon pumps identified in the previous sections: the solubility pump, the organic carbon pump and the calcium carbonate counter pump. Several feedback processes can be identified that are expected to operate on these pumping mechanisms to a greater or lesser extent on the timescales of interest to climate prediction (decades to centuries).

4.1 Temperature - Carbonate Chemistry

As the ocean warms, CO_2 solubility decreases and the carbonate equilibrium shifts in favour of higher concentration of carbonic acid (due to the effect of temperature on the apparent equilibrium constants). These two effects combine to increase pCO_2 in the ocean by 4-5%/°C for fixed values of alkalinity and total carbon (Lashof, 1989). This reduces the uptake of atmospheric CO_2 into the mixed layer, resulting in a higher atmospheric pCO_2 and an increased warming due to the greenhouse effect.

4.2 Temperature - Vertical Mixing

Higher surface layer temperatures could result in an increased stability of the thermocline and a reduction in vertical eddy diffusion. This would result in a slower rate of uptake of both CO_2 and heat by the deep ocean. Both effects act so as to raise surface temperatures. Reduced vertical mixing and increased heat storage in the mixed layer would also lead to a decrease in the supply of nutrients to the surface, resulting in a reduction in the efficiency of the biological pump due to the increased separation between sunlight and nutrient supply.

4.3 Temperature - Meridional Circulation

From the few equilibrium CO_2 experiments performed so far with fully coupled atmosphere-ocean GCMs, it appears that a weakening of the mean meridional circulation of the Atlantic occurs when CO_2 is doubled (Manabe *et al*, 1990), owing to increased stratification at high latitudes caused by increases in

surface temperature and precipitation. If these predictions are correct then this would result in a reduction of non-biological CO_2 drawdown at high latitudes (ie a reduction in the solubility pump), leaving higher concentrations of CO_2 in the atmosphere and thereby enhancing the greenhouse effect.

4.4 Wind Distribution - Gas-exchange Rate

The gas transfer coefficient governing the air-sea exchange of CO_2 increases with wind speed. Recent measurements of gas exchange under storm conditions (Watson *et al*, 1991) indicate a steep increase in transfer velocity once the wind reaches such a strength that the sea surface begins to break up with whitecaps. A feedback between CO_2-induced climate change and CO_2 uptake by the ocean then becomes possible due to changes in global wind speeds and storm frequency. Thus if a significant shift in the global wind pattern were to occur it could influence the rate of gas transfer at the sea surface. However, equilibrium between the atmosphere and the mixed layer is realised over a time scale of less than a year, whereas exchange between the mixed layer and sub-surface waters is usually much slower than this. Thus in most locations the net rate of air-sea exchange does not constrain the overall uptake of CO_2 by the oceans and any feedback effect is likely to be quite modest. Indeed, carbon cycle models show that the net uptake of CO_2 is not sensitive to the gas transfer coefficients used, since uptake is controlled mainly by vertical mixing rather than through gas exchange (Oeschger *et al*, 1975).

4.5 Biological Pump Feedbacks

Studies using simple box models indicate that the solubility pump accounts for only about 30% of the partitioning of CO_2 between the mixed layer and deep sea (see, e.g. Volk and Hoffert, 1985). Hence any changes to the biological pump should be of great importance for climate studies.

A number of possible feedback mechanisms exist for ocean biology. Enhanced levels of primary production would increase the drawdown of CO_2, and reduce surface and hence atmospheric values. Thus global warming could be reduced or enhanced according to whether a warmer climate resulted in increased or decreased levels of production respectively. Increased temperature is unlikely to have any great direct effect on productivity if changes occur gradually over a long enough period of time. Similarly, increased concentrations of CO_2 will have little direct effect since primary production tends to be limited by the availability of nutrients and/or light rather than by the supply of CO_2. However, a change in the relative abundance of carbon compared with the abundance of limiting nutrients could have important consequences through changes in the efficiency of the biological cycling of carbon.

An absolute increase in the nutrient supply could increase drawdown of CO_2. Assuming that photosynthesis was still nutrient-limited at the sea surface so that concentrations were still low there, the vertical nutrient gradient would be enhanced, since all available nutrients would still be used up at the surface whilst higher concentrations would occur below the euphotic zone. This would lead to a higher rate of transfer of nutrients to the sea surface and a concomitant increase in the rate of photosynthesis. With this would be an associated increased detritus flux into deeper layers. If the extra nutrients were associated with CO_2-induced warming this would result in a negative

feedback. An increased nutrient supply could be caused by a rise in sea level due to greenhouse warming - this would increase the area of continental shelf which is much more productive than the open ocean. Alternatively, rising sea levels might also lead to the deposition of nutrient-rich organic sediments, causing a loss of nutrients from the ocean (Broecker, 1982) and lower productivity, but since the deposition of shelf sediments is a relatively slow process compared with the timescales of human-induced climate change it does not concern us here. Another negative feedback is that due to a reduction in the extent of sea ice found in polar upwelling areas, leading to an increased penetration of light, sustaining higher productivity (since phytoplankton growth is thought to be limited by light rather than by nutrients in these regions).

Probably the most significant biological feedback mechanism so far identified involves a reorganisation of the atmosphere-ocean circulation system and consequent changes in ocean productivity. Changing patterns of circulation and mixing may serve to enhance or reduce global productivity and may have either an inhibitory or stimulatory effect on the transport of carbon into the deep ocean. Such changes might occur through variations in the wind forcing, resulting in changed upwelling patterns, or by changes in vertical mixing, bottom water formation etc. The magnitude and direction of such a feedback is at present unknown, but is thought to have been important in changes that occurred during the last deglaciation. Primary production in polar and subpolar regions is thus thought to be a key factor in the maintenance of lower atmospheric pCO_2. Changes in the strength and location of such regions of high production could affect the distribution of marine ecosystems and species composition, possibly affecting pCO_2 in surface waters. An example of this would be a shift from diatoms (which use silica to form cell walls and don't affect alkalinity) to coccolithophores (which use calcium carbonate and hence reduce alkalinity) as the dominant species. In regions where this occurs it would decrease surface

alkalinity, leading to reduced ocean uptake of CO_2 by the alkalinity pump feedback (see later). The global significance of such changes in species composition is not known. Toggweiler and Sarmiento (1985), using a three-box model, have shown that the high latitude outcrop regions play a central role in controlling atmospheric pCO_2 through the rate of nutrient and carbon supply by deep convection relative to the biological uptake of these constituents and their export to the deep ocean. These processes might in principle change quite rapidly. A time-dependent solution of this three-box model (Wenk and Siegenthaler, 1985) shows that after an instantaneous perturbation of one of the controlling parameters it takes of the order of 200 years to reach a new equilibrium.

In equatorial waters, recent data (Barber, 1990) indicate that relatively modest increases in upwelling-favourable winds can cause a 40% increase in equatorial productivity. This suggests a global warming-productivity feedback whereby warming leads to a reduced meridional temperature gradient and a reduction in the thermal wind (ie reduced wind shear). If this produced weaker surface winds it would cause weaker upwelling and reduced productivity, leading to a reduced downward transfer of carbon, increased surface pCO_2, an increase in atmospheric pCO_2 and further greenhouse warming. However, reduced upwelling will simultaneously reduce degassing from the upwelled carbon-rich water, making the net atmospheric effect of a reduction in upwelling at the equator and along the continental margins difficult to predict.

Another mechanism involving primarily low latitude regions is the 'Alkalinity Pump' (Sarmiento *et al*, 1988). It concerns a change in the relative efficiency of the organic and inorganic components of the biological pump. Dissolution of $CaCO_3$ occurs at significantly greater depths than the recycling of organic carbon, with the consequence that alkalinity is driven deeper down into the water

column than ΣCO_2. Most organic carbon is recycled before it falls below the depth of the thermocline, whereas a greater proportion of the inorganic particle flux extends into deeper water. Hence higher productivity and/or reduced exchange between deep water and the thermocline, relative to the exchange between the surface mixed layer and the thermocline, could give rise to a large flux of calcium carbonate to the deep sea. This would have the net effect of reducing upper ocean alkalinity by trapping it in the deep ocean (alkalinity is a conservative quantity and is conserved globally). Since pCO_2 varies inversely with surface alkalinity, an increase in the deep ocean sequestering efficiency, or more vigorous surface mixing (relative to the deep sea), would lead to an increase in surface ocean, and hence atmospheric, partial pressure of CO_2. There is, however, a natural limit to this feedback. An atmospheric pCO_2 beyond 1100 ppmv is unlikely due to the fact that the calcareous organisms that drive this process would have difficulty surviving in the undersaturated waters that develop at this point. This process may have played a role in the decreased pCO_2 of the ice ages.

7. Conclusion

A better understanding of the global carbon cycle is of central importance in predicting global change, and the oceans represent a crucial component. The physical, biological and chemical processes affecting CO_2 are highly interactive and several feedback mechanisms may be important. In order to predict reliably the rate at which the oceans take up CO_2 from the atmosphere, ocean models are required that are realistic enough to simulate the relevant processes that occur, including water circulation and mixing, the chemical and biological transformations of carbon, and the consequent movement of chemical tracers in the ocean.

Until recently the only carbon cycle models available have been relatively simple box models which have been calibrated by calculating fluxes of water between boxes and diffusion coefficients within boxes so that observed tracer distributions are reproduced in the models. Such methods assume an unchanging ocean circulation and are not suitable for predicting CO_2 distribution into the far future in a system in which numerous feedbacks occur. Furthermore, the relative importance of many of the feedbacks so far identified has not as yet been established. For these purposes a 3D circulation model is required in which carbon is transported around the atmosphere-ocean system and the ocean is allowed to respond to changing atmospheric forcing. This would require a model of the carbon cycle that uses the advective velocities computed by the ocean circulation model and includes equations describing the chemical and biological transformations of the relevant chemical constituents: ΣCO_2, alkalinity, limiting nutrient concentration, dissolved oxygen, particulate and dissolved organic carbon. It may be necessary to explicitly include the annual phytoplankton bloom that occurs outside tropical waters, particularly if the biological flux of carbon out of the mixed layer proves to be susceptible to changing seasonal patterns of forcing and nutrient supply. Ideally, such a chemical/biological model would be incorporated into an eddy-resolving ocean model to allow for biases due to the patchy growth of phytoplankton. Such an approach is not yet feasible with present day computers. However, early results from coarse resolution, non-eddy-resolving, models are quite promising and show good agreement between model predictions of phytoplankton blooms and satellite photographs. It may thus be possible to explicitly model the complex three-dimensional physics and the feedback mechanisms described earlier without going to the expense of an eddy-resolving model. The challenge is to provide a carbon cycle model that is detailed enough to include the most important mechanisms and feedbacks, but which is also as simple and economic as possible so that it can be incorporated into global coupled atmosphere-ocean general circulation models.

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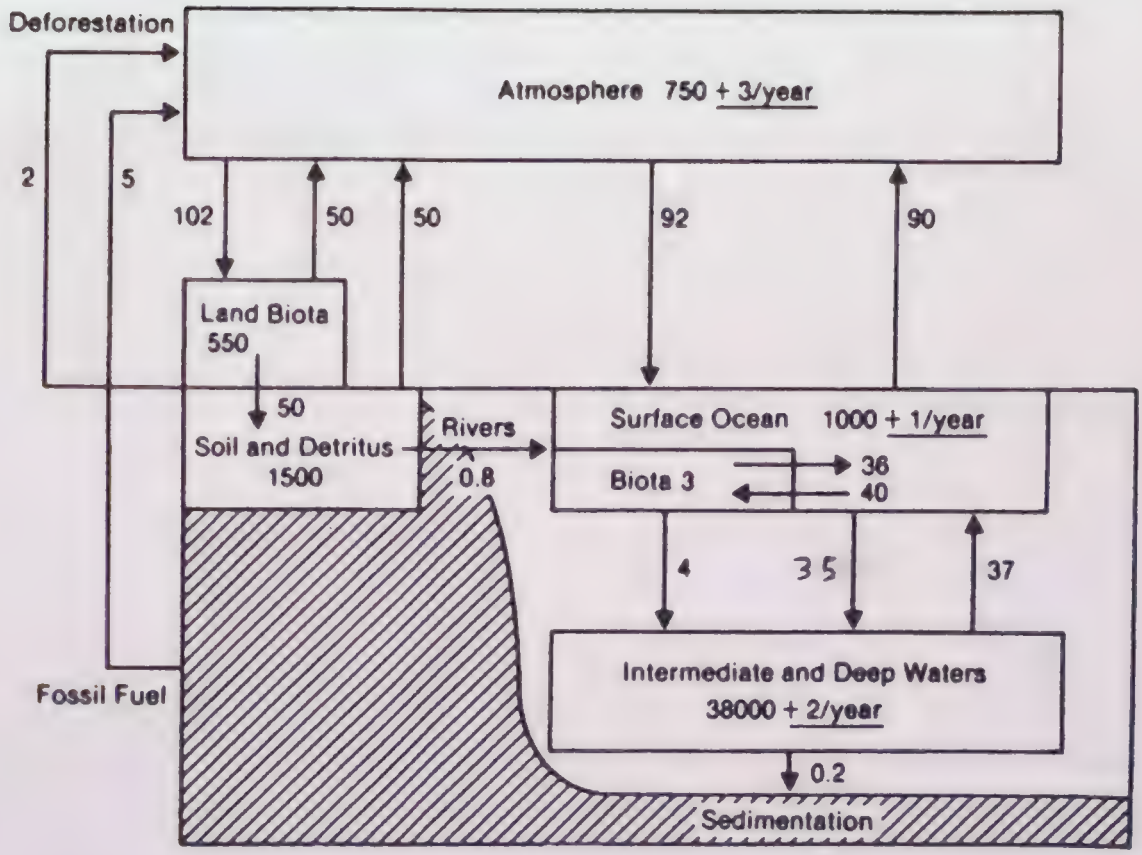

Figure 1 Global carbon reservoirs (in Gt C) and gross annual fluxes (in Gt C yr^{-1}). Numbers underlined indicate net annual CO_2 accumulation due to human action. (From IPCC, 1990)

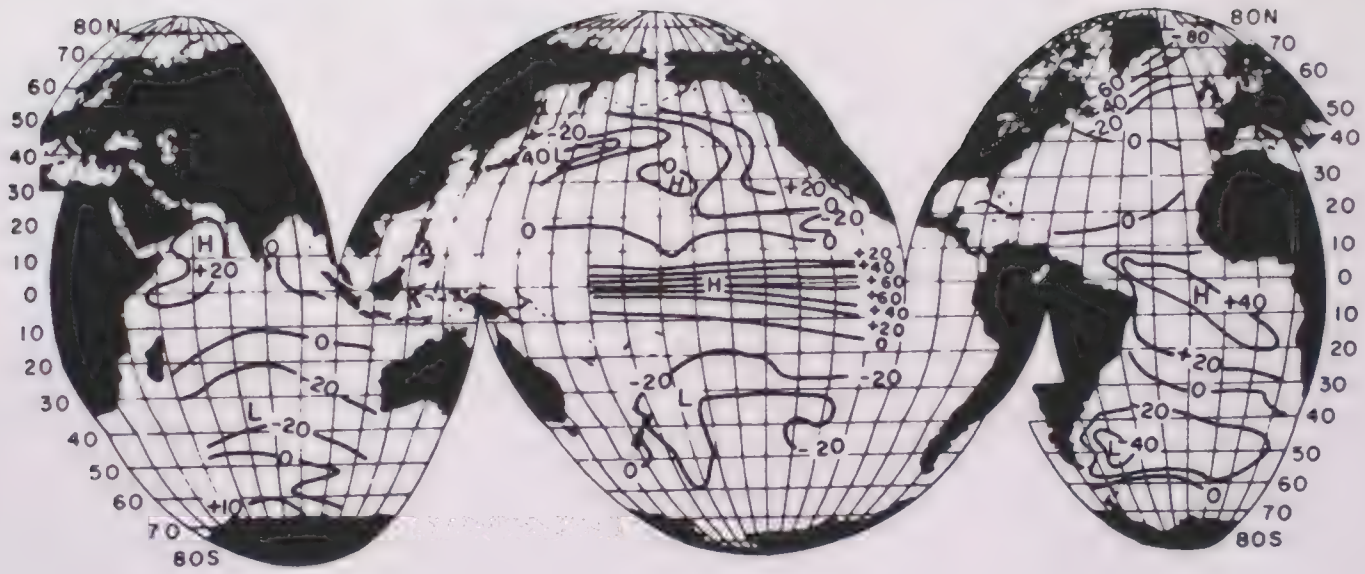

Figure 2 Global distribution of the difference [pCO_2 (surface waters)–pCO_2 (atmosphere)] in microatmospheres observed during the GEOSECS program. The data represent summer conditions in both hemispheres. (From Takahashi and Azevado, 1982)

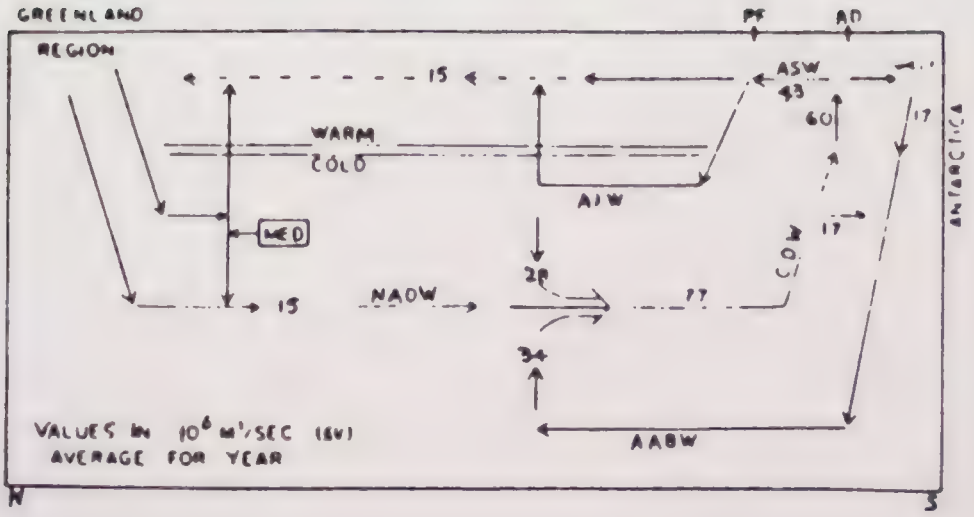

Figure 3 Schematic view of the Atlantic meridional circulation according to Gordon (1971). NADW=North Atlantic Deep Water, ASW=Antarctic Surface water, AIW=Antarctic Intermediate Water, AABW=Antarctic Bottom Water, CDW=Combined Deep Water.

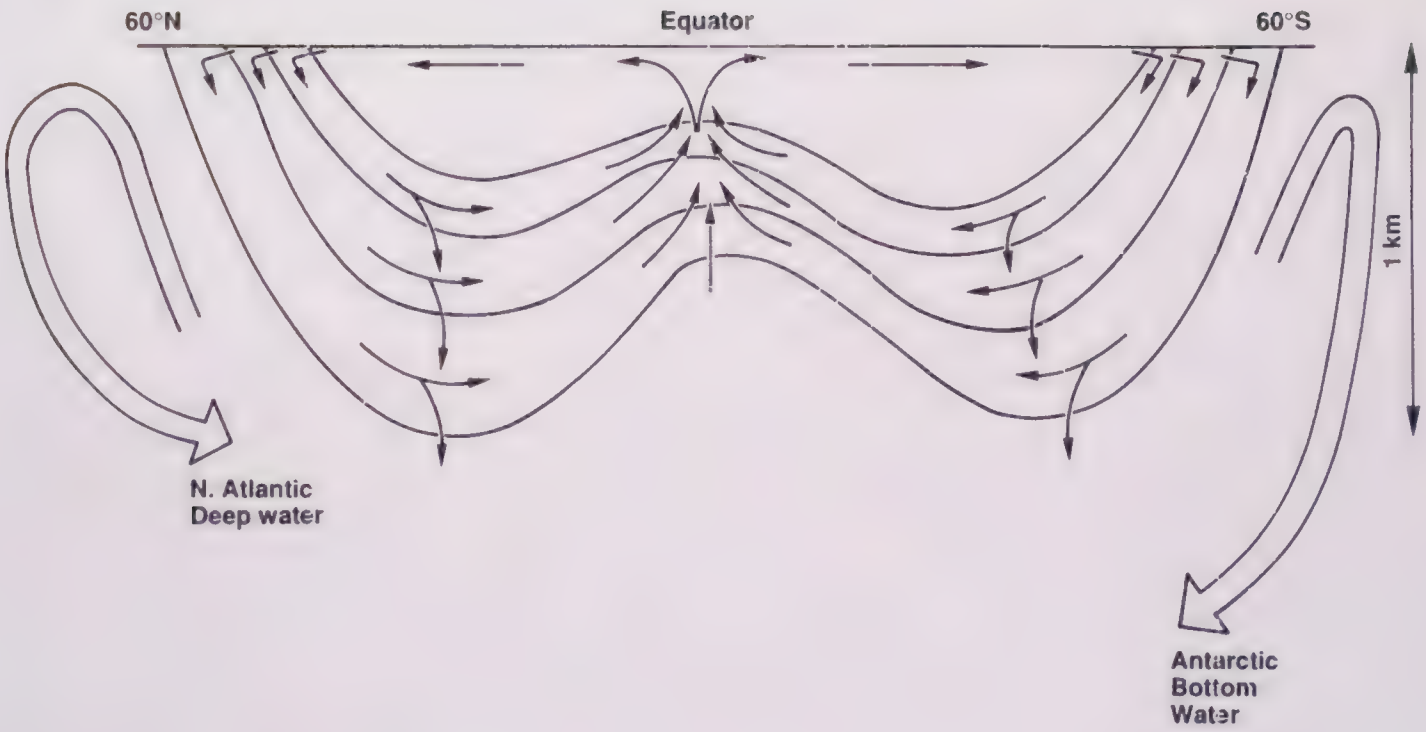

Figure 4 Schematic meridional section of the Atlantic, showing the pathways of water ventilating the thermocline and the formation of deep water at high latitudes.

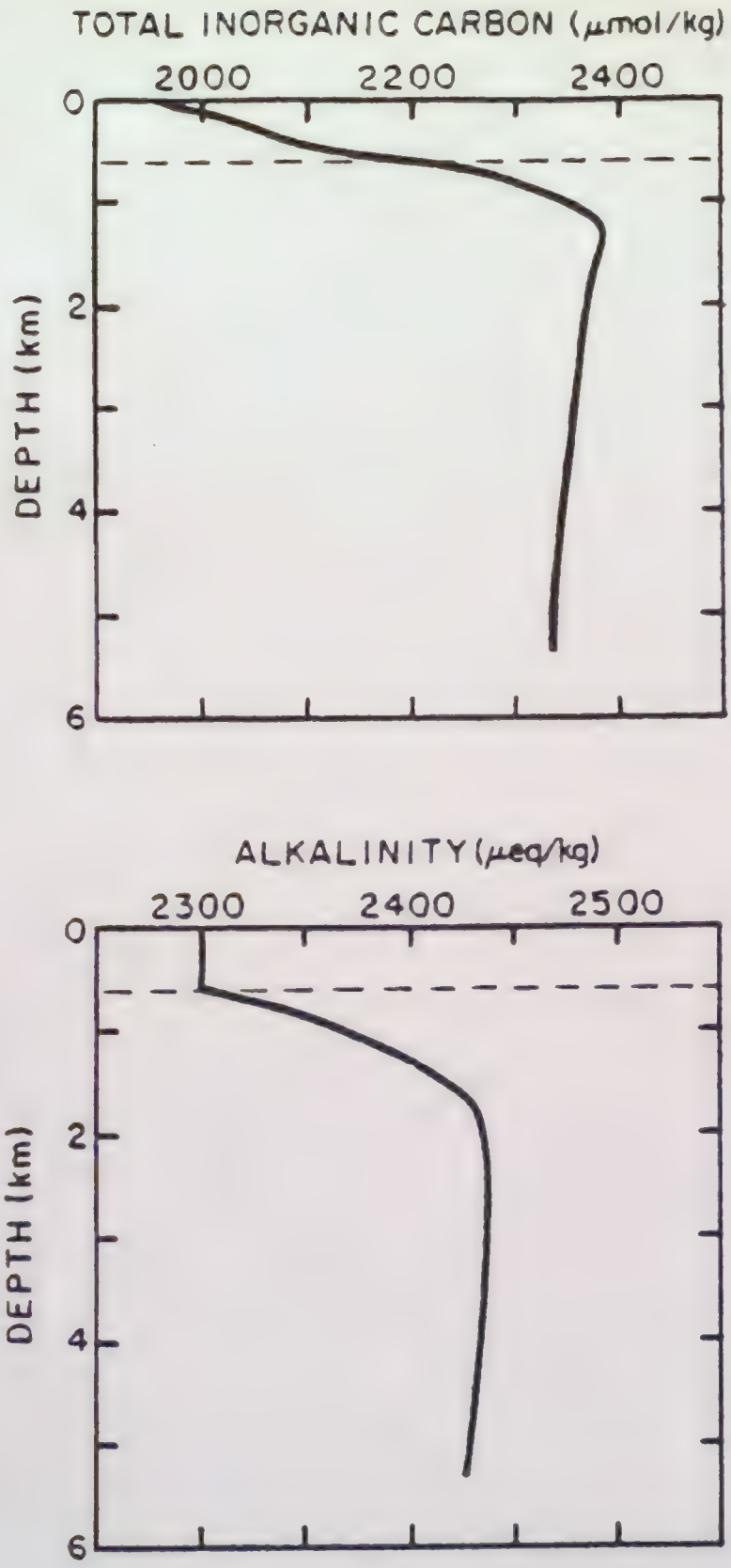

Figure 5 Depth distribution of dissolved inorganic carbon and alkalinity at Geosecs station 214 in the North Pacific (32°N,176°W). (From Broecker and Peng, 1990).

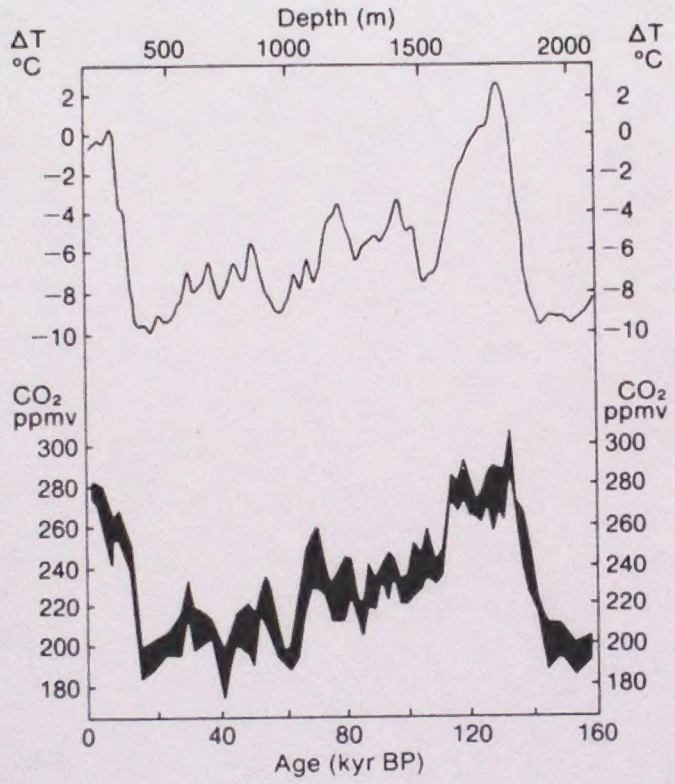

Figure 6 CO_2 concentrations (bottom) and estimated temperature changes (top) during the past 160,000 years, as determined from an Antarctic ice core. Atmospheric CO_2 concentration and temperature show a high degree of correlation, with rapid variations occurring over periods of a few centuries. (From IPCC, 1990)

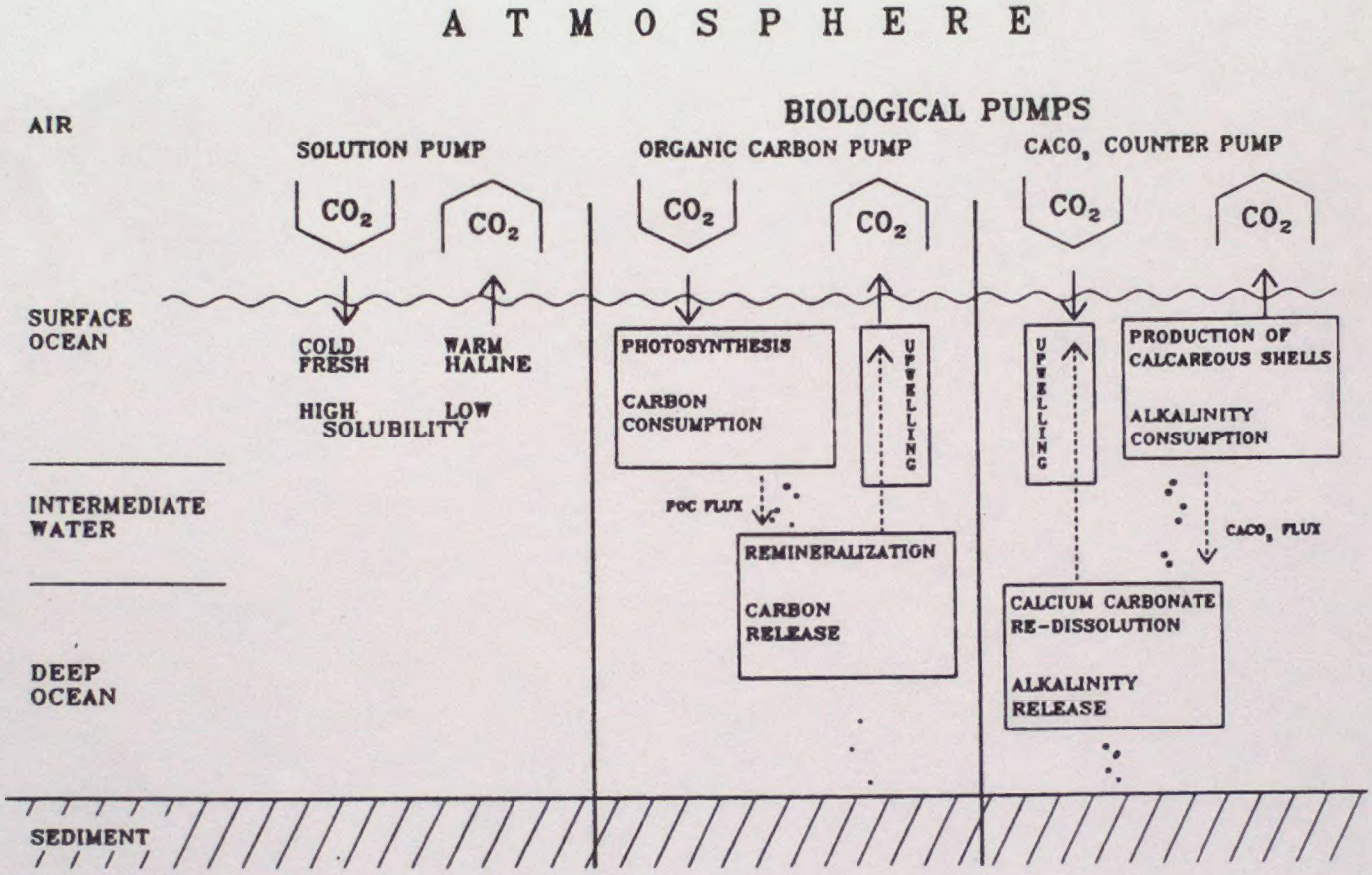

Figure 7 Schematic diagram of the marine carbon pumps. (From Heinze *et al*, 1990).

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